GEOPHYSICAL RESEARCH LETTERS, VOL. 27, NO. 22, PAGES 3623-3626, NOVEMBER 15, 2000

The Thermal State of the Upper Mantle; No Role for Mantle Plumes

Don L. Anderson
Seismological Laboratory, California Institute of Technology, Pasadena

Abstract. A variety of geophysical data indicates that long wavelength temperature variations of the asthenosphere depart from the mean by ± 200°C, not the ± 20°C adopted by plume theoreticians. The 'normal' variation, caused by plate tectonic processes (subduction cooling, continental insulation, small-scale convection) encompasses the temperature excesses that have been attributed to hot jets and thermal plumes. Geophysical estimates of the average potential temperature of the upper mantle are about 1400°C. Asthenospheric convection at ridges, rifts and fracture zones and at the onset of continental breakup is intrinsically 3D, giving rise to shallow pseudo-plume-like structures without deep thermal instabilities. Deep narrow thermal plumes are unnecessary and are precluded by uplift and subsidence data. The locations and volumes of 'midplate' volcanism appear to be controlled by lithospheric architecture, stress and cracks.

Introduction

Verhoogen (1973), Elder (1976) and Richter (1973) pointed out the importance of lateral temperature gradients and small scale convection in the upper mantle. More recently, theoretical models of terrestrial magmatism assume that the normal state of asthenospheric mantle is isothermal, and thus static, and subsolidus. This applies to both passive (e.g. Boutilier and Keen, 1999) and plume (e.g. White and McKenzie, 1989) models. However, systems cooled from above or having lateral temperature, conductivity or radioactivity gradients at the top will develop small-scale convection. This topside driven convection can be an order of magnitude faster than plate rates or rifting rates (Korenaga, 2000) and cannot be ignored or treated as a small perturbation. Rapid convection increases the melt delivery to the surface in regions of extension without an increase in mantle temperature. The plume hypothesis, to a large extent, is based on the hypothesis that the 'normal' state of the mantle is isothermal, cold, subsolidus, static, refractory, dry and homogeneous, and that 3D, focusing and small-scale convection effects are not important. The locations of volcanoes and the variations in crustal thickness depend on more than temperature. The ability of rift-induced dynamic convection to explain large igneous provinces from 'normal' temperature mantle obviates the need for hot deep mantle plumes (King and Anderson, 1998; Boutilier and Keen, 1999). Topside tectonics also explains plate motions and continental breakup (Elder, 1976; Lowman and Jarvis, 1999).

McKenzie and Bickle (1988) (M & B) assumed the upper mantle to be homogeneous and more-or-less isothermal. They adopted a 'cold' subsolidus potential temperature of 1280°C ± 20°C and assumed that temperatures that are 150° to 200°C higher than this are due to localized hot jets. This type of isothermal mantle implies either no convection, or very high (> 10^{12}) Rayleigh number convection (e.g., Niemela et al., 2000).

Geophysical estimates of upper mantle temperatures are more than 100°C hotter (e.g. Anderson and Bass, 1984; Anderson, 1989; Kaula, 1983; Hofmeister, 1999), and the normal variability is about 10 times greater than assumed by M & B. There is very little support for the cold isothermal asthenosphere hypothesis, a corollary of the plume hypothesis. If normal mantle temperatures are 1400° ± 200°C, there is no thermal requirement for plumes. Small scale convection associated with ridges, rifts and edges is intrinsically 3D, giving rise to concentrated, shallow, plume-like upwellings (Richter, 1973; Parmentier and Phipps Morgan, 1990). Cracks and dikes may control the dimensions of volcanic features. Thus, there is also no geometric requirement for the deep mantle plume hypothesis. Geophysical and petrological estimates of mantle potential temperatures (the P = 0 extension of the adiabat) are summarized in this paper. A shallow source for geochemical variations was addressed earlier (Anderson, 1994, 1995).

Temperature Variations in the Upper Mantle

The range of temperatures in the upper mantle, below the thermal boundary layer ('fully convective region' in the nomenclature of Kaula (1983)) is easier to constrain than the absolute temperature. Bathymetry, rate of seafloor subsidence, heat flow, and depths of the 410 and 650 km discontinuities are all functions of temperature and complement the standard petrological approaches utilizing crustal chemistry and thickness variations. There is a remarkable consistency in estimates derived from these various datasets (Table 1).

Kaula (1983) estimated the minimal upper mantle temperature variations that are consistent with observed heat flow and plate velocities. At the fully convective level, about 280 km depth, temperature variations are at least ± 180°C, averaged over 500 km spatial dimensions. This is in contrast to the assumption of M & B that at this depth all geotherms are horizontal.

Temperature variations along the global midocean ridge system are about 200°C (Klein and Langmuir, 1987; Kane and Hayes, 1994; Perrot et al., 1998). This represents about half the global range. The composition of near-ridge peridotites implies a temperature range of 200°C (Bonatti, 1990). Some so-called hotspots are actually wet-spots of normal temperature. In particular, the Azores platform appears to be related to near-ridge fracture zones and transform faults and is not underlain by hot mantle (Azevedo and Portugal, 1999).

The thickness of the transition region (TR) (400 to 650 km depth) constrains the mantle temperature and the temperature variation (Anderson, 1967; 1989). Flanagan and Shearer (1998) obtain 244 ± 32 km as the thickness. This gives a tem-

Paper number 2000GL011533.
0094-8276/00/2000GL011533$05.00

Table 1. Long Wavelength Temperature Variations in the Sublithospheric Mantle

GLOBAL	
Plate velocities (Kaula, 1983)	± 180°C
Petrology (Klein and Langmuir, 1987)	± 125°C
Subsidence (Calcagno and Cazenave, 1993)	± 100°C
Transition Zone thickness	± 100°C
Lower Mantle (Yoneda and Spetzler, 1994)	± 112°C
North America (Butler, 1984)	± 145°C
Theoretical (Anderson, 1998a)	± 200°C
COLD REGIONS	
Cratons (Li et al., 1998)	< -150°C
Subduction (Anderson, 1998a)	- 150°C
Deep ridges (Bonatti et al., 1994; Lanyon et al., 1995)	- 150°C
Non-ridges (Kaula, 1983)	- 100°C to -250°C
HOT REGIONS	
Supercontinent insulation (Anderson, 1998a)	+ 200°C
Ridges (Kaula, 1983)	+ 100° to + 250°C
Iceland (Sato and Sacks, 1989)	~ + 120°C
Iceland (Ribe et al., 1995)	< + 70°C
"Hotspots" (Ribe et al., 1995)	+ 50° to + 70°C
(White and McKenzie, 1989)	+ 150° to + 200°C
(Skogseid et al., 1992)	+ 50° to + 130°C
(Ito and Lin, 1995a)	+ 50° to + 150°C
(Schilling, 1991)	+ 162° to 278°C

perature variation of ± 120°C to ± 230°C depending on choice of thermochemical parameters (Anderson, 1989; Agee, 1998). A recent study gives 250 ± 10 km as the global range (Chevrot et al., 1999) which gives a ΔT of ± 100°C in TR, consistent with Li et al. (1998). Melbourne and Helmberger (2000) determined that TR thicknesses under the entire East Pacific Rise (EPR) are indistinguishable from the global mean (PREM; Dziewonski and Anderson (1981)). This region contains three or four proposed hotspots yet they do not influence TR temperatures, suggesting shallow roots, not only for ridges, but also for regions of excess volcanism.

The Chevrot et al. (1999) study shows no TR thinning under Iceland, Hawaii, Easter, Afar, Yellowstone or Cameroons. The shallow mantle in some of these regions, and also the EPR, has low seismic velocities; high temperatures and large temperature fluctuations apparently do not extend to 650 km. Regional studies near subduction zones imply TR temperatures 200-300°C colder than average (Clarke et al., 1995).

It can be noted that a temperature rise of 200°C can bring an upper mantle rock from subsolidus to one that is 20% molten. Although the deepest and coldest parts of the global ridge system are about 175°C colder than average, they still provide basalts (Bonatti et al., 1993, 1994; Lanyon et al., 1995). This suggests that "average ridge" mantle is above the solidus.

The above estimates of temperature fluctuations are consistent with the ± 10% variability which accompanies "normal" convection (Elder, 1976; Lowman and Gable, 1999; Niemela et al., 2000) and are of the order required to drive 3D shallow mantle small scale convection and plume-like instabilities (Davaille and Jaupart, 1994). These shallow plume-like instabilities can deliver even larger volumes of melt from normal temperature mantle than 2D rolls (Korenaga, 2000) or hot deep mantle plumes (Cordery et al., 1997).

Absolute Temperature

The absolute temperature of the mantle is harder to constrain than the temperature variation, but values estimated from mineral physics, seismology, geodynamics, heat flow and petrology are consistent (Kaula, 1983; Anderson and Bass, 1984; Duffy and Anderson, 1989; Anderson, 1989). These approaches give a mean potential temperature of about 1400°C for the upper mantle. Equation of state fits to the lower mantle yield potential temperatures of 1500°C (Zhao and Anderson, 1994; Stacey, 1992) to 1730°C (Stixrude et al., 1992). The average conductive geotherm (Hofmeister, 1999) intersects the upper mantle adiabat slightly below 80 km depth, and the wet peridotite or eclogite solidus at shallower depth. In the warmer parts of the mantle the geotherm may be on or above the solidus to depths of 300 km (Anderson and Bass, 1984; Anderson, 1989). These studies indicate that colder parts of the mantle are below the solidus throughout.

Global tomographic studies show that ridges and hotspots occur over broad regions of hotter than average mantle (Anderson et al., 1992; Wen and Anderson, 1997). If the average temperature of the mantle is close to the melting point, the inference is that the ridge and hotspot mantle is near the solidus to depths greater than 250 km.

Convection driven by edge effects and variable lithospheric thickness can deliver melts at the volumes and rates required at large igneous provinces from mantle with temperatures in the 'normal' range (King and Anderson, 1998; Boutilier and Keen, 1999). Shallow plume-like 3D effects are even more effective (Korenaga, 2000), and likewise do not require substantial heating from below.

Hotspot Temperatures

Plumes are a hypothetical form of convection based on the premise that excess magmatism require localized regions of high temperature. Since normal convection and plate tectonic processes can cause variations of ± 200°C and excess magmatism can also be caused by small-scale convection and magma focusing, there is no *à priori* need for deep mantle plumes. Geophysical estimates of hotspot temperatures are in the range of "normal" mantle temperatures (Table 1).

Ribe et al. (1995) and Feighner et al. (1995) show that hotspots are much colder than once thought (e.g. Schilling, 1991). They derive temperature 'excesses' of <70°C for the Azores, Galápagos, and Iceland. Ito and Lin (1995b) used bathymetry and gravity to estimate near-ridge temperatures adjacent to hotspots. The temperature excesses are generally between 50° and 150°C. Temperatures of Galápagos basalts (1186°C ± 30°C) are less than along the nearby ridge segment (Fisk et al., 1982).

Schilling (1991) attempted to infer excess temperatures of plumes from bathymetry. Values for 13 hotspots fall in the narrow range of + 162°C (for Tristan) to + 278°C (for Circe), much higher than later estimates. There is no relation to hotspot-ridge distance or discharge rate, as expected from plume theory. These are upper bound temperature estimates.

Local variations in bathymetry and subsidence rates of oceanic lithosphere imply temperature variations of 100° to 200°C in regions picked to avoid 'hotspot influence' (Kane and Hayes, 1994). Thus, the temperature 'excesses' attributed to ridge-hotspot interactions are within the range of normal mantle temperature variations. The much larger excess temperatures required by the plume hypothesis (e.g. Cordery et al., 1997) are not supported by the data. Iceland, Azores, Tristan, Galápagos and Easter are five hotspots with prominent bathymetric and geochemical anomalies, yet these imply temperature anomalies of less than 150°C (Ito and Lin, 1995b). They can be regarded as near-ridge fracture zone and edge phenomena and regions of small-scale convection and lithospheric extension (Sykes, 1978; Richter, 1973; Anderson, 1998b).

Mantle temperatures can also be inferred from guyot heights (Caplan-Auerbach et al., 2000). Most seamounts and volcanic islands were emplaced on seafloor overlying mantle less than 100°C hotter than average. The median thermal anomaly for the Easter and Emperor chains is 150°C. Guyots along the Hawaiian, Marquesas, Louisville, Marshall and Darwin Rise imply temperature excesses of less than 100-200°C.

Takahashi et al. (1998) argues that ocean island basalts are from fertile mantle rather than hot mantle, and that most estimates and assumptions of ocean island and continental flood basalt temperatures and volumes are too high. They calculate that hotspots originate in mantle of 1400°C or less.

Korenaga (2000) determined that the mantle temperature associated with the breakup of Greenland from Europe and the North Atlantic igneous province was 1270° - 1350°C throughout the rifting process and that changes in the volumes of extrusives were related to small-scale convection. The crust of the Kolbeinsey Ridge, just north of Iceland, implies a constant temperature of 1320-1360°C for the past 22 Ma (Kodaira et al., 1998). This hotspot province is well within the range of 'normal' mantle temperatures.

Seismic investigations of Hawaii indicate a deep cold root extending to at least 100 km (Woods and Okal, 1996; Priestley and Tilman, 1999). The Woods and Okal (1996) study shows that the upper 200 km under Hawaii is similar to normal Pacific mantle. Inferred temperatures are relatively low (1463°C), and the top of the melt zone is deeper than in the plume calculations of Watson and McKenzie (1991). There is no evidence from seismology that the transition zone under Hawaii is thin (hot) (Chevrot et al., 1999). Multiple ScS phases, bouncing under Hawaii, and the Galápagos, show normal, or fast, mantle velocities (Best et al., 1974; Sipkin and Jordan, 1976). Some factor other than temperature, such as lithospheric architecture and stress, is controlling the locations and volumes of midplate volcanism. Most volcanoes of all kinds occur above upper mantle that is seismically slow, or hot, but specific locations appear to be controlled by the lithosphere.

Causes of Lateral Temperature Gradients

Normal Bénard convection, without narrow plumes, implies variations for the mantle of ± 200°C (Niemela et al., 2000). Temperature variations at the top of the convecting mantle are also caused by slab cooling, cratonic roots and continental insulation (Anderson, 1998a). These are imposed lateral temperature changes in contrast to the accidental or induced gradients set up by Bénard convection driven by bottom heating. The imposed variations amount to ± 200°C and have various characteristic length scales. The shears imposed by moving plates and rise of the asthenosphere between cratons impose other length scales and generate another dimension of motion (horizontal rolls and vertical stalks) into what appear to be 2D problems (linear ridges and rifts). Lateral temperature gradients alone can induce 3D upwellings. The plume hypothesis ignores these effects and the resultant shallow plume-like upwellings and attributes all such features to deep thermals.

Summary

The geophysical data that constrains the lateral variations of temperature below the plate include: bathymetry, subsidence rates, heat flow, global plate motion modeling, depths to mantle phase changes, seismic velocities, thickness of the transition region and crustal thickness. These data imply temperature variations of ± 150° to ± 200°C, even when filtered to avoid 'hotspot influence' and subduction zones. There is good agreement between various geophysical estimates of 'normal' upper mantle temperature variations.

The potential temperature of the upper mantle is 1400° ± 200°C based on long-wavelength bathymetry, subsidence, heat flow, tomography, plate motions, discontinuity depths and petrology. The mean is more than 100°C hotter than assumed for 'normal' mantle in the plume hypothesis.

The absence of appreciable thermal anomalies associated with continental flood basalts (Czamanske et al., 1998; Korenaga, 2000) suggests that rapid fluxing of the asthenosphere through the melting zone and 3D effects (Richter, 1973; Parmentier and Phipps Morgan, 1990) are responsible for excess magmatism, not hot mantle plumes. "Topside tectonics" is now a more mature and self-consistent theory than plume theory and does not require an *ad hoc* initial singularity (e.g. Cordery et al., 1997) to get it started.

Acknowledgments. This paper represents Contribution Number 8702, Division of Geological and Planetary Sciences, California Institute of Technology. This work has been supported by NSF Grant EAR 9726252.

References

Agee, C. B., Phase transformations and seismic structure in the upper mantle and transition zone in *Rev. Mm.*, ed. R. J. Hemley and P. H. Ribbe, vol. 37, pp. 165-203, Min. Soc. Am., Washington D. C., 1998.

Anderson, D. L., Phase changes in the upper mantle, *Science, 157*, 1165-1173, 1967.

Anderson, D. L., *Theory of the Earth*, Blackwell Scientific Publications, Boston, 366 p., 1989.

Anderson, D. L., The sublithospheric mantle as the source of continental flood basalts, *Earth Planet. Sci. Lett., 123*, 269-280, 1994.

Anderson, D. L., Lithosphere, asthenosphere and perisphere, *Rev. Geophys., 33*, 125-149, 1995.

Anderson, D. L., The scales of mantle convection, *Tectonophysics, 284*, 1-17, 1998a.

Anderson, D. L., The EDGES of the mantle, in *The Core-Mantle Boundary Region*, ed. M. Gurnis et al., pp. 255-271, Am. Geophys. U., Washington D.C., 1998b.

Anderson, D. L., and J. D. Bass, Mineralogy and composition of the upper mantle, *Geophys. Res. Lett., 11*, 637-640, 1984.

Anderson, D. L., T. Tanimoto, and Y.-S. Zhang, Plate tectonics and hotspots: the third dimension, *Science, 256*, 1645-1651, 1992.

Azevedo, J. M. M., and M. R. Portugal, Volcanic gaps and subaerial records of palaeo-sea-levels on Flores Island, *J. Geodyn., 28*, (2-3), 117-129, 1999.

Best, W. J., L. R. Johnson, and T. V. McEvilly, ScS and the mantle beneath Hawaii, *Eos, 55*, 1147, 1974.

Bonatti, E., et al., A cold suboceanic mantle belt at the Earth's equator, *Science, 261*, (5119), 315-320, 1993.

Boutilier, R. R., and C. E. Keen, Small-scale convection and divergent plate boundaries, *J. Geophys. Res.*, *104*, (B3), 7389-7403, 1999.

Calcagno, P., and A. Cazenave, Present and past regional ridge segmentation, *Geophys. Res. Lett.*, *20*, (18), 1895-1898, 1993.

Caplan-Auerbach, J., F. Duennebier, and G. Ito, Origin of intraplate volcanoes from guyot heights and oceanic paleodepth, *J. Geophys. Res.*, *105*, (B2), 2679-2697, 2000.

Chevrot, S., L. Vinnik, and J.-P. Montagner, Global-scale analysis of the mantle Pds phases, *J. Geophys. Res.*, *104*, (9), 20,203-20,219, 1999.

Clarke, T. J., et al., Close in ScS and SScS reverberations from the 9 June, 1994 Bolivian Earthquake, *Geophys. Res. Lett.*, *22*, (16), 2313-2316, 1995.

Cordery, M. J., et al., Genesis of flood basalts from eclogite-bearing mantle plumes, *J. Geophys. Res.*, *102*, 20179-20197, 1997.

Czamanske, G. K., et al., Demise of the Siberian plume, *Int. Geol. Rev.*, *40*, 95-115, 1998.

Davaille, A., and C. Jaupart, Onset of thermal convection in fluids with temperature-dependent viscosity, *J. Geophys. Res.*, *99*, 19853-19866, 1994.

Duffy, T. S., and D. L. Anderson, Seismic velocities in mantle minerals and the mineralogy of the upper mantle, *J. Geophys. Res.*, *94*, 1895-1912, 1989.

Duffy, T. S., and T. J. Ahrens, The effect of temperature on V_P at high pressure; implications for mantle dynamics, *Eos*, *73*, 317, 1992.

Dziewonski, A. M., and D. L. Anderson, Preliminary Reference Earth Model, *Phys. Earth Planet. Int.*, *25*, 297-356, 1981.

Elder, J., *The Bowels of the Earth*, p. 222, Oxford University Press, London, 1976.

Feighner, M., et al., Numerical modeling of chemically buoyant mantle plumes at spreading ridges, *Geophys. Res. Lett.*, *22*, 715-718, 1995.

Fisk, M. R., A. E. Bence, and J. G. Schilling, Major element chemistry of Galapagos rift zone magmas and their phenocrysts, *Earth Planet. Sci. Lett.*, *61*, (1), 171-189, 1982.

Flanagan, M. P, and P. M. Shearer, Global mapping of topography on transition zone velocity discontinuities by stacking SS precursors, *J. Geophys. Res.*, *103*, (2), 2673-2692, 1998.

Hofmeister, A. M., Mantle values of thermal conductivity and the geotherm from phonon lifetimes, *Science*, *283*, (5408), 1699-1706, 1999.

Ito, G. T., and J. Lin, Mantle temperature anomalies along the present and Paleo-axes of the Galapagos Spreading Center as inferred from gravity analyses, *J. Geophys. Res.*, *100*, (B3), 3733-3745, 1995a.

Ito, G. T., and J. Lin, Oceanic spreading center-hotspot interactions, *Geology*, *23*, (7), 657-660, 1995b.

Kane, K. A., and D. E. Hayes, A new relationship between subsidence rate and zero-age depth, *J. Geophys. Res.*, *99*, (11), 21,759-21,777, 1994.

Kaula, W. M, Minimal upper mantle temperature variations consistent with observed heat flow and plate velocities, *J. Geophys. Res.*, *88*, 10323-10332, 1983.

King, S. D., and D. L. Anderson, Edge-driven convection, *Earth Planet. Sci. Lett.*, *160*, 289-296, 1998.

Klein, E. M., and C. H. Langmuir, Global correlations of ocean ridge basalt chemistry with axial depth and crustal thickness, *J. Geophys. Res.*, *92*, (8), 8089-8115, 1987.

Kodaira, S., et al., Evolution of oceanic crust on the Kolbeinsey Ridge, *Terra Nova*, *10*, (1), 27-31, 1998.

Korenaga, J., Magmatism and Dynamics of Continental Breakup in the Presence of a Mantle Plume, Ph.D. Thesis, MIT, 2000.

Lanyon, R., A. J. Crawford, and S. M. Eggins, Westward migration of Pacific Ocean upper mantle into the Southern Ocean region between Australia and Antarctica, *Geology*, *23*, (6), 511-514, 1995.

Li, A., et al., Mantle discontinuities and temperature under the North American continental keel, *Nature*, *395*, (6698), 160-163, 1998.

Lowman, J. P., and C. W. Gable, Thermal evolution of the mantle following continental aggregation in three-dimensional convection models, *Geophys. Res. Lett.*, *26*, (17), 2649-2652, 1999.

Lowman, J. P., and G. T. Jarvis, Effects of mantle heat source distribution on supercontinent stability, *J. Geophys. Res.*, *104*, (6), 12,733-12,747, 1999.

McKenzie, D., and M. J. Bickle, The volume and composition of melt generated by extension of the lithosphere, *J. Petrol.*, *29*, (3), 625-679, 1988.

Melbourne, T. and D. Helmberger, Precise transition zone thickness estimates from triplicated multiple S, manuscript, January 2000.

Niemela, J. J., et al., Turbulent convection at very high Rayleigh numbers, *Nature*, *404*, 837-840, 2000.

Parmentier, E. M. and J. Phipps Morgan, Spreading rate dependence of three-dimensional structure in oceanic spreading centres, *Nature*, *348*, 325-328, 1990.

Perrot, K., et al., Spatial and temporal variations of subsidence of the East Pacific Rise, *Earth Planet. Sci. Lett.*, *160*, (3-4), 593-607, 1998.

Priestley, K., and F. Tilman, Shear-wave structure of the lithosphere above the Hawaiian hot spot, *Geophys. Res. Lett.*, *26*, (10), 1493-1496, 1999.

Ribe, N. M., et al., The dynamics of plume-ridge interactions; Ridge-centered plumes, *Earth Planet. Sci. Lett.*, *134*, (1-2), 155-168, 1995.

Richter, F. M., Dynamic models for sea floor spreading, *Rev. Geophys.*, *11*, 223-287, 1973.

Sato, H., and I. S. Sacks, Constraint on magma generation in island arc region from seismic Q studies, *Bull. Volcan. Soc. Japan*, *34*, (2), 148, 1989.

Schilling, J. G., Fluxes and excess temperatures of mantle plumes inferred from their interaction with migrating midocean ridges, *Nature*, *352*, (6334), 397-403, 1991.

Sipkin, S. A., and T. H. Jordan, Lateral heterogeneity of the upper mantle determined from the travel times of multiple ScS, *J. Geophys. Res.*, *81*, (35), 6307-6320, 1976.

Skogseid, J., et al., Tectonism and magmatism during NE Atlantic continental break-up, in *Magmatism and the causes of continental break-up*, ed. B. C. Storey, et al., vol. 68, p. 305-320, Geolo. Soc. Spec. Pub., London, 1992.

Stacey, F. D., *Physics of the Earth*, p. 513, Brookfield Press, Kenmore (Australia), 1992.

Stixrude, L., et al., Thermodynamic analysis of new high P-T equation of state data and implications for the composition of the lower mantle, *Intl. Geol. Congress* - Japan, *29*, p. 55, 1992.

Sykes, L. R., Intraplate seismicity, reactivation of pre-existing zones of weakness, alkaline magmatism, and other tectonism postdating continental fragmentation, *Rev. Geophys. Space Phys.*, *16*, (4), 621-688, 1978.

Takahashi, E., et al., Origin of the Columbia River basalts, *Earth Planet. Sci. Lett.*, *162*, (1-4), 63-80, 1998.

Verhoogen, J., Possible temperatures in the oceanic upper mantle and the formation of magma, *GSA Bull.*, *84*, 515-522, 1973.

Watson, S., and D. McKenzie, Melt generation by plumes, European Union of Geosciences, *Terra Abstracts*, *3*, (1), 418, 1991.

Wen, L., and D. L. Anderson, Slabs, hotspots, cratons and mantle convection revealed from residual seismic tomography in the upper mantle, *Phys. Earth Planet. Inter.*, *99*, 131-143, 1997.

White, R., and D. McKenzie, Magmatism at rift zones; the generation of volcanic continental margins and flood basalts, *J. Geophys. Res.*, *94*, (6), 7685-7729, 1989.

Woods, M. T., and E. A. Okal, Rayleigh-wave dispersion along the Hawaiian Swell, *Geophys. J. Intl.*, *125*, (2), 325-339, 1996.

Yoneda, A., and H. Spetzler, Temperature fluctuation and thermodynamic properties in Earth's lower mantle, *Earth Planet. Sci. Lett.*, *126*, (4), 369-377, 1994.

Zhao, Y., and D. L. Anderson, Mineral physics constraints on the chemical composition of the Earth's lower mantle, *Phys. Earth Planet. Inter.*, *85*, 273-292, 1994.

D. L. Anderson, California Institute of Technology, Seismological Laboratory 252-21, Pasadena, CA 91125 (e-mail: dla@gps.caltech.edu)

(Received February 24, 2000; revised May 18, 2000; accepted September 29, 2000.)